CONCLUSIONS

1. A mechanical distribution of gages such as described in this study appears to give reasonably thorough sampling of rainfall variation in mountain watersheds. Additions to or modifications of such distributions in order to improve sampling may be made by a study of isohyetal maps and statistical constants based upon a preliminary rain gage installation.

2. The gage system employed iu this experiment gives results accurate for most storms measured, within practical limits.

3. In order to avoid employing an impracticably large number of rain gages, the requirements for accuracy of averages should be modified in inverse relation to the size and importance of storms.

4. With a system of gages distributed so as to sample rainfall variation as thoroughly as possible, a simple average of their readings will agree within close limits with rainfall catch computed from isohyetal maps. Application of the former method requires much less time and skill than the isohyetal method.

PRELIMINARY RESULTS OF PILOT-BALLOON ASCENTS AT LITTLE AMERICA

By G. GRIMMINGER

[Weather Bureau, Washington, D. C., June 1939]

The meteorological observations at Little America during the two Byrd Expeditions include 969 pilot-balloon ascents. These have been worked up fairly completely and the results of the two years of observation combined, although it has not yet been possible to give the results the study they deserve, it will be of interest to point out some of the more obvious facts which they disclose.

From the combined data for both the expeditions, there have been computed the mean direction and mean velocity as well as the resultant direction, resultant velocity, and stability. The above quantities have also been worked out for each of the 4 seasons.

The mean direction should be distinguished from the resultant direction; the former is computed by using the frequency with which the different directions occur, while the latter is computed by using the velocities as well as the directions and gives the direction of the vector representing the resultant air transport.

The stability, mentioned above, refers to directional stability, and is computed by forming the ratio $\frac{\text{resultant velocity}}{\text{mean velocity}}\times 100$; this gives a measure of the steadiness of the wind direction, for, if the wind always blows from the same direction, the resultant velocity and the mean velocity are the same and the stability is 100 percent. If, on the other hand, the directions are equally distributed and also have the same velocity, the resultant velocity and therefore the stability will be zero.

TABLE 1.—*Mean values of direction, velocity, and stability of the wind at Little America* [1]

[Based on 2 years of observation, 1929 and 1934]

Altitude (meters)	Number of observations	Mean direction from—	Resultant direction from—	Mean velocity	Resultant velocity	Stability	Mean direction—resultant direction
				m. p. s.	*m. p. s.*	*Percent*	
12,000	4	N 19° W	N 7° W	17.5	6.8	39	−12°
11,000	9	N 8° W	N 18° W	11.6	5.7	49	10°
10,000	30	N 8° E	N 5° W	8.9	3.2	36	13°
9,000	65	N 12° E	N 14° W	11.0	2.2	20	26°
8,000	115	N 10° W	N 59° W	13.0	3.7	28	49°
7,000	172	N 23° W	N 66° W	12.5	3.2	26	43°
6,000	236	N 37° W	N 70° W	11.2	2.4	21	33°
5,000	318	N 55° W	S 85° W	9.7	1.2	12	40°
4,000	415	S 1° E	S 28° W	8.4	1.0	12	−29°
3,000	549	S	S 29° W	7.7	1.4	18	−29°
2,500	625	S 9° E	S	7.4	1.5	20	−9°
2,000	704	S 17° E	S 9° E	7.4	2.0	27	−8°
1,500	772	S 18° E	S 9° E	7.3	2.2	30	−9°
1,000	854	S 20° E	S 10° E	7.4	2.0	27	−10°
750	883	S 25° E	S 16° E	7.5	2.0	27	−9°
500	924	S 34° E	S 30° E	7.7	2.3	30	−4°
250	957	S 35° E	S 40° E	7.1	2.9	41	5°
Surface	969	S 22° E	S 40° E	4.0	1.8	45	18°

[1] Latitude 78°34′06″ south, longitude 163°55′58″ west.

In table 1 are given the mean values of the direction, velocity, and stability of the wind at standard levels for the 2 years of observation, 1929 and 1934. The vertical

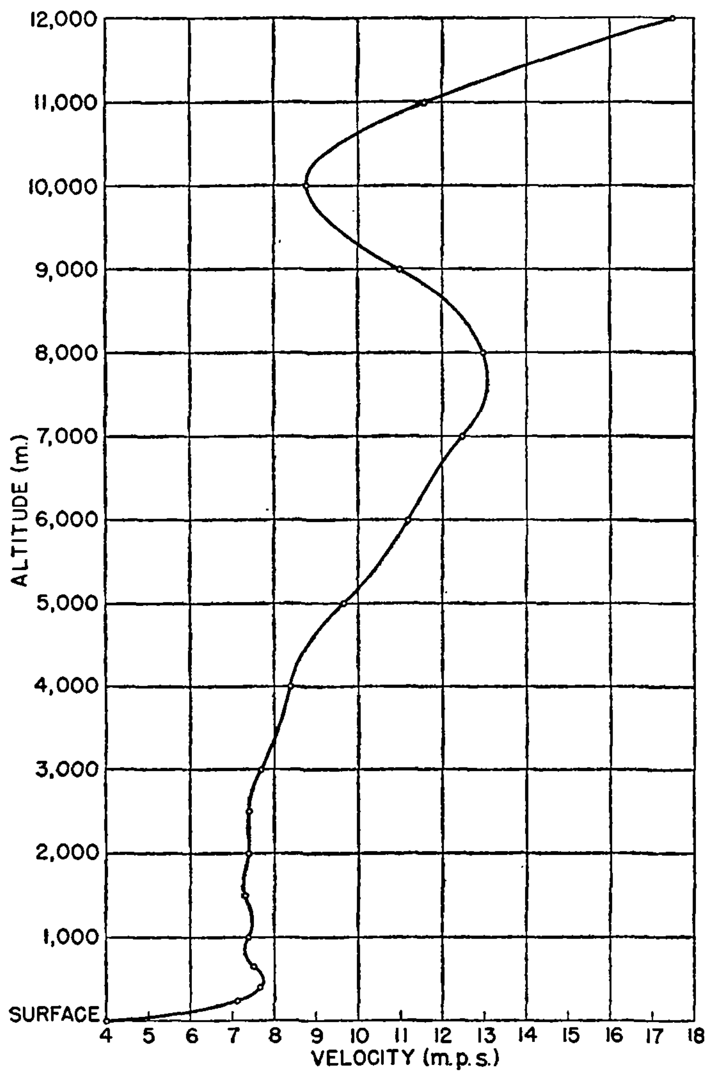

FIGURE 1.—Vertical distribution of the mean wind velocity at Little America.

distribution of the mean velocity is shown in figure 1, from which it is seen that a maximum value is reached at about 7,700 m. From the studies of Peppler (1) and Dobson (1A), this wind maximum can be related to the height of the tropopause. From the results of his study Peppler gave the following table.

Height of tropopause	8	9	10	11	12	13	14 km.
Height of wind maximum	8.0	8.7	9.3	10.0	10.6	11.3	12.0 km.
Difference	0.0	0.3	0.7	1.0	1.4	1.7	2.0 km.

In the region in which the observations were made there is thus a relation between the height of the wind maximum and the height of the tropopause, such that the two practically coincide for low values of the tropopause height.

Although Peppler's study was based on data for Central Europe, there seems to be no reason why a similar relation should not hold in general for the other latitudes as well; and in the polar regions, where there is much evidence pointing to a relatively low tropopause, one may reasonably expect the height of the wind maximum and that of the tropopause to be more or less coincident. Thus the mean height of the tropopause for the year can be considered as equal to, or perhaps slightly greater than, 7,700 m.

The only other available upper air-wind data in the Antarctic consist of 120 pilot-balloon ascents made during the drift of the ship *Deutschland* in the Weddel Sea during the German Antarctic Expedition in 1911–12. The results of these ascents as given by Barkow (**2**) show a maximum wind velocity at about 7,500 m., which is in very good agreement with the value of 7,700 m. found at Little America. In the North Polar region we have the observations made by Sverdrup (**3**) during the Maud Expedition, which show a velocity maximum at 8,000 m., which is in good agreement with the above values for the southern hemisphere.

The separate years, 1929 and 1934, at Little America likewise show good agreement, the average height of the velocity maximum being about 7,900 m. in 1929, and 7, 500 m. in 1934. There is a well-marked variation in height of the tropopause from summer to winter of about 1,100 m., the tropopause being higher in summer and lower in winter, which agrees with results of temperature soundings in other latitudes.

The above results do not agree with the estimate by Shaw (**4**) who, in a chart showing the distribution of the tropopause over the globe, placed this below 4 km in winter over the south polar area. As stated above, on the basis of the wind observations, the height of the tropopause over the Antarctic is essentially the same as over the Arctic, and Shaw's estimate appears to be about 3.5 km too low.

The increase in velocity indicated in figure 1 at 11 and 12 km is possibly not representative of average conditions, but rather the result of the very small number of observations at these levels.

Some insight into the reason for the velocity maximum at or near the tropopause can be gained by a consideration of the effect of horizontal gradients of temperature upon the variation of wind with height. Letting the y-axis point north and the x-axis east, and assuming geostropic conditions, the following equations are easily derived (*5*):

$$\frac{\partial u}{\partial z}=\frac{1}{\rho f T}\left(\frac{\partial p}{\partial y}\frac{\partial T}{\partial z}-\frac{\partial p}{\partial z}\frac{\partial T}{\partial y}\right) \quad (1)$$

$$\frac{\partial v}{\partial z}=\frac{-1}{\rho f T}\left(\frac{\partial p}{\partial x}\frac{\partial T}{\partial z}-\frac{\partial p}{\partial z}\frac{\partial T}{\partial x}\right) \quad (2)$$

where u is the East-West component of the velocity and v the North-South component.

Assuming as a first approximation that in sufficiently high layers of the atmosphere, conditions are symmetric with respect to longitude then $\frac{\partial p}{\partial x}$, $\frac{\partial T}{\partial x}$, and consequently $\frac{\partial v}{\partial z}$, are zero, and we need consider only $\frac{\partial u}{\partial z}$. From equation (1) we can now investigate the sign of $\frac{\partial u}{\partial z}$ in the upper troposphere. Using mean values of the horizontal gradients of pressure and temperature along a meridional cross section[1], and mean values of the vertical gradients of pressure and temperature, it is found that in the upper troposphere $\frac{\partial u}{\partial z}$ is positive. In the stratosphere $\frac{\partial T}{\partial z}$ can be taken as zero and since the horizontal temperature gradient is reversed, $\frac{\partial u}{\partial z}$ becomes negative. Thus, at or near the tropopause $\frac{\partial u}{\partial z}$ changes sign from positive to negative which means that u must have a maximum value at this altitude. The wind maximum is thus essentially the result of the reversed temperature gradient in the stratosphere.

The result of combining all the upper air wind data for the 2 years of observation at Little America shows that the average wind direction is southeasterly in the lower levels and undergoes a clockwise rotation with increasing altitude, thus becoming in turn southeasterly, southernly, southwesterly, westerly, northwesterly, and northerly; the northerly winds occurring at the highest levels. This turning of the wind direction is indicated by both the mean direction and the resultant direction.

There is a fairly well-marked minimum in the stability at 4 and 5 km., indicating that the greatest variability in wind direction occurs at these levels; this is also the altitude where the *mean* wind direction changes from one having a southerly component to one having a northerly component. It thus appears that on the average there is a transition layer between 4 and 5 km. separating the troposphere into two parts.

Thus in the lower part of the troposphere there is a south component of the resultant wind which is fairly constant up to about 1,500 m. and then decreases gradually to zero at about 5 km. At all levels above this there is a north component which gradually increases with elevation up to 10 km., which is probably as far as the computations can be considered as significant owing to the small number of observations at the higher levels.

In the last column of table 1 is given the difference between the mean direction and the resultant direction, the positive sign indicating that the mean direction is in advance of the resultant direction in the sense of a clockwise rotation. One might expect that the greatest difference between the mean and the resultant direction would occur where the stability is a minimum. This apparently is not the case, since the greatest difference occurs at 8 km., that is, at the approximate level of the tropopause. The minimum stability on the other hand seems to be related to the change in sign of the difference between the two directions. In this particular case the effect of the change in signs is such that the level of the minimum stability coincides with the level where the change of the direction difference with altitude has its maximum value, namely, 69° per km.

[1] See *Handbuch der Klimatologie*, Band 1, Teil F., S. 66, 67.

Considering the east-west components of the resultant wind, there is an east component in the lower layers which gradually decreases with elevation and becomes zero at 2,500 m. Above this there is a west component at all levels, and this has a maximum at around 8 km. corresponding to the altitude of the maximum mean wind.

As a result of the decrease of temperature southward in the upper layers, there will also be a decrease of pressure so that the westerly winds are to be associated with a polar cyclone. Or, looking at this in another way, if the wind conditions in the upper layers are considered as existing more or less symmetrically around the continent, then the westerly winds indicate the presence of a polar cyclone in the upper levels.

The question now arises as to what layers are to be included in the cyclonic circulation. If we consider this to consist of all those layers having a westerly component of motion, the polar cyclone will then begin at about 2,500 m. and extend to 10 km. or higher. Assuming the resultant wind to be geostrophic, the center of the cyclone for the layer between 2.5 and 5 km. will be situated somewhere over West Antarctica; from 5 to 8 km. the center will be more or less centered over the polar plateau itself, while above 8 km. the center will be over East Antarctica corresponding to the large northerly component in the motion.

The easterly component of motion in the lower layers has usually been interpreted as being the result of a polar anticyclone; so that since the east component in the yearly average vanishes at 2,500 m., the depth of the anticyclone cannot be greater than this.

In the past, since pilot-balloon data were lacking, it was customary to assume that the average direction of the winds at upper levels was essentially the same as that derived from the observed directions of cloud movement. In order to test the validity of such an assumption, table 2, based on 2 years of observation at Little America, has been prepared.

TABLE 2.—*Comparison of mean direction of cloud motion and mean direction of wind at Little America*

	Lower clouds (St.)	Number of observations	Intermediate clouds (A St)	Number of observations	Upper clouds (Ci St)	Number of observations
Mean height of clouds, meters.	892	157	3,031	72	5,070	15
Mean direction of clouds.[1]	N. 50° E.	557	N. 58° E.	278	N. 7° W.	134
Mean direction of wind.	S. 23° E.	854	S.	549	N. 55° W.	318
Difference	107°		122°		−48°	
Resultant direction of wind.	S. 13° E.	854	S. 29° W.	549	S. 85° W.	318
Difference	117°		151°		−88°	

[1] From surface observations.

The mean cloud heights were determined from the pilot-balloon ascents; the table contains the differences: mean cloud direction minus mean wind direction; and mean cloud direction minus resultant wind direction. It is seen that these differences are large, amounting to 122° and 151°, respectively, for the level of the intermediate clouds at 3,000 m, and 48° and 88° respectively for the level of the upper clouds at 5,000 m. These results certainly seem to indicate that any deduction of the average direction of air flow from observations of cloud motion may be greatly in error, especially when the mean cloud altitude is near or within the levels of minimum stability. This is not surprising since conditions producing an overcast of ASt or CiSt can hardly be considered as representing the average. It thus seems, especially at certain levels, that the mean direction of the isobars when taken to be the same as the mean direction of cloud motion can be very much in error.

The polar anticyclone must be manifested by winds having an easterly component, and it is reasonable to associate greater intensity of this anticyclone with greater easterly wind components. In table 3 are given the east-west components in summer, winter, and the three coldest months; and it is seen that in the lower layers the east components are much smaller in winter than in summer, while for the three coldest months it is found that the resultant wind direction has no east component at all. This is hardly what one would expect, since, as a result of the intense radiational cooling in the winter, it would appear that the anticyclonic circulation should be intensified rather than weakened. In fact the common explanation of the polar anticyclone is that the cooling of the air lowers the isobaric surfaces and the resulting inflow aloft raises the surface preasure (over what is was before) thus leading to increased anticyclonic circulation. Incidentally this explanation is hardly consistent with the fact that the sea-level pressure, at least in the Ross Sea area, is distinctly lower in the coldest months.

TABLE 3.—*East-west components of the resultant wind velocity at Little America*

Altitude (meters)	Summer (December, January, and February)	Winter (June, July, and August)	3 coldest months (July, August, and September)
	m. p. s.	*m. p. s.*	*m. p. s.*
9,000	0.72		3.49
8,000	4.68		2.43
7,000	3.24	3.87	3.69
6,000	1.97	4.42	2.87
5,000	1.43	2.07	1.35
4,000	.75	.96	1.84
3,000	−.19	.97	1.73
2,500	−.72	.92	1.75
2,000	−1.00	.22	1.54
1,500	−1.15	.00	1.54
1,000	−.87	.12	1.44
750	−1.22	−.10	1.35
500	−1.94	−.37	.99
250	−2.40	−.78	.71
Surf	−1.47	−.43	.37
East = − West = +			

That the decrease in the east component in the cold season is real, and not the result of having made the balloon ascents during particular weather types, is substantiated by the results of the hourly surface-wind observations which are given in the following table:

Mean surface wind direction

	Summer	Winter	3 coldest months
From hourly values	S. 50° E.	S. 11° E.	S. 7° W.
From pilot-balloon ascents	S. 26° E.	S. 5° E.	S. 16° W.

The mean direction as computed from the hourly values shows the decrease in the east component from summer to winter and also the absence of any east component during the 3 coldest months. For the sake of comparison there is given in the second row the mean surface direction as computed from the pilot-balloon ascents and it is seen that these show substantial agreement with the results of the hourly values indicating that the results of the pilot-balloon ascents may be considered as giving a satisfactory representation of average conditions.

It will be noted that in the upper layers the west components are larger in winter from 4 to 7 km, and in the 3 coldest months from 4 to 9 km, excepting at 8 km, where a weakening is indicated.

Since most of the Antarctic Continent is in the form of an elevated land mass with a downward slope from the Pole to the edges of the continent, radiational cooling in the lower layers must lead to a drainage of cold air outward and as a result of the deflective influence of the earth's rotation this drainage effect alone would act to increase the east component of the atmospheric motion in the lower layers. The outward drainage to the north in the lower layers will lead to an inflow aloft toward the south which, as a result of the deflective influence, would be indicated by northwesterly winds in the upper levels. As stated below the *annual* means show very clearly winds with south and east components at the low levels and with north and west components at the higher levels and agrees qualitatively with the concept of the outward drainage of cold air below and a compensating inflow aloft. During the colder months, one should therefore expect an increase in the south component in the lower layers and increased north components aloft. In table 4 are given the north-south components of the resultant wind velocity in summer winter, and the 3 coldest months. It is seen that the south components are distinctly larger in winter than in summer up to 1,000 m; this is even more pronounced during the coldest period and extends to 2,000 m. In the upper layers it is seen that there are larger north components in winter than in summer at 7 and 8 km. and in the coldest period from 5 to 9 km.

TABLE 4.—*North-south components of the resultant wind velocity at Little America*

Altitude (meters)	Summer (December, January, and February	Winter (June, July, and August)	3 coldest months (July, August, and September)
	m. p. s.	*m. p. s.*	*m. p. s.*
9,000	3.81	----------	3.80
8,000	3.16	4.17	3.47
7,000	1.47	2.79	1.58
6,000	1.20	.68	1.46
5,000	.49	−.21	1.07
4,000	−.95	−.29	−.04
3,000	−1.31	−.27	−.73
2,500	−1.69	−.52	−1.36
2,000	−2.14	−1.35	−2.19
1,500	−2.24	−2.04	−2.67
1,000	−1.61	−2.59	−3.46
750	−1.53	−2.96	−3.74
500	−1.53	−2.96	−3.74
250	−1.80	−3.03	−3.57
Surface	−1.34	−1.31	−1.50
South = − North = +			

In general, the seasonal variation of the north-south components agrees with what would be expected if the circulation over the continent were controlled mainly by the drainage of the cold lower layers. The seasonal variation of the east-west components in the upper layers also agrees with this; but that of the lower layers does not, since as pointed out above the east components are smaller in winter and are lacking entirely during the 3 coldest months. In high latitudes where the Coriolis effect attains its maximum value, any drainage of the cold air toward the north in the lower layers should be acted on by the deflecting influence to a maximum extent so as to produce southeast or east winds. Just why these east components do not appear in the coldest months when the drainage should be a maximum is not clear.

It should be emphasized, however, that the direction of the drainage flow must depend largely on the direction of slope of the surface. Little America is situated on the Ross Shelf Ice, which is a large area of low, level surface surrounded by high land to the west, east, and south, that has a general downward slope into the shelf ice area. Thus, there is South Victoria Land to the west, Marie Byrd land to the east, and the Queen Maud Range and Polar Plateau to the south. This area is therefore well situated to act as a drainage basin for the cold air from these high slopes and to serve as an outlet for it to the north. This is supported by the fact of the larger south component in the air transport found in the coldest months. Along the western side of the shelf ice the damming effect of the high South Victoria land is sufficient to prevent any easterly component in the movement which would arise as a result of the earth's rotation and to cause the main motion there to take place in a north-south direction. It is doubtful, however, if this effect would be present at Little America since it is 400 or more miles to the east.

It should also be pointed out that easterly winds can be produced by cyclonic as well as anticyclonic pressure distributions, and it would be important to separate, if possible, the contribution of these two types of pressure distribution in producing easterly winds. It is also important to study the relation between the eastward moving depressions and the release of large masses of cold air.

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TROPICAL DISTURBANCE OF JUNE 12–16, 1939

By JEAN H. GALLENNE

[Marine Division, Weather Bureau, August 1939]

The first tropical disturbance of 1939 attained only moderate intensity, but moved in a rather unusual course from the Gulf of Honduras northward and north-northwestward to the east Gulf coast.

The earliest report of disturbed conditions in connection with this depression was received on the morning of June 12, through the Mexican weather service at Chetumal, placing the center near latitude 18°45′ N. and longitude 87° W. During the afternoon of the same day, although no reports of strong winds were received, vessels in the area just east of the Yucatan Peninsula reported squally weather conditions, with moderate to rough seas.